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Review Article

Moho and magmatic underplating in continental lithosphere

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ABSTRACT

Underplating was originally proposed as the process of magma ponding at the base of the crust and was inferred from petrologic considerations. This process not only may add high density material to the deep crust, but also may contribute low density material to the upper parts of the crust by magma fractionation during cooling and solidification in the lower crust. Separation of the low density material from the high-density residue may be a main process of formation of continental crust with its characteristic low average density, also during the early evolution of the Earth. Despite the assumed importance of underplating processes and associated fractionation, the available geophysical images of underplated material remain relatively sparse and confined to specific tectonic environments. Direct ponding of magma at the Moho is only observed in very few locations, probably because magma usually interacts with the surrounding crustal rocks which leads to smearing of geophysical signals from the underplated material. In terms of processes, there is no direct discriminator between the traditional concept of underplated material and lower crustal magmatic intrusions in the form of batholiths and sill-like features, and in the current review we consider both these phenomena as underplating. In this broad sense, underplating is observed in a variety of tectonic settings, including island arcs, wide extensional continental areas, rift zones, continental margins and palaeo-suture zones in Precambrian crust. We review the structural styles of magma underplating as observed by seismic imaging and discuss these first order observations in relation to the Moho.

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Contents

1. Introduction	0
2. Underplating and lower crustal reflectivity	0
3. Compressional settings	0
3.1. Arc magmatism and crustal formation	0
3.2. Underplating in Precambrian crust	0
4. Extensional settings	0
4.1. Wide extensional areas	0
4.2. Large batholiths and sills in a Moho transition zone	0
4.3. Rift zones: magma compensated crustal thinning	0
4.3.1. Modern rifts	0
4.3.2. Palaeo-rifts	0
4.4. Volcanic rifted continental margins	0
5. Large igneous provinces, plumes, and cratons	0
6. Discussion	0
7. Conclusions	0
Acknowledgments	0
References	0

1. Introduction

The oldest crust was created from mantle derived magma by processes similar to underplating, which formed the original Moho. The formation of continental crust requires multiple melting sequences

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to form magma that, after solidification, will have the characteristic low average density of the crystalline basement (Arndt, [this volume](#); Hawkesworth et al., [this volume](#)). The continental crust is subsequently affected by a variety of tectonic, erosional, depositional and metamorphic processes, which define the evolution of individual regions. Clearly, plate boundary processes, at both subduction and rift zones, play major roles in the shaping of the continental crust by tectonic and magmatic processes. These processes and their importance for forming the continental lithosphere have been widely discussed with focus on the various tectonic regimes. Additionally, the interaction of mantle upwelling (plumes) with continental lithosphere may play an important role in lithosphere growth, modification, and destruction both at plate margins and in intraplate regions. Mantle melting and infiltration of basaltic magmas are not restricted to the mantle part of the lithosphere, but often result in emplacement of magmatic bodies into the crust or at its base, i.e. as underplated material. In the following, we mainly focus on intraplate settings and the effects of the interaction of tectonic and magmatic processes that lead to intrusion of magma around the Moho.

Underplating and intrusion of magma into the crust and uppermost mantle are important processes for crustal formation and subsequent evolution because the addition of magma provides a non-tectonic way for the crust to grow and thicken. The resulting solidified magma bodies remain in the crust and uppermost mantle as images of past processes spanning the whole sequence of tectomagmatic processes, including crustal formation, orogenesis, rifting and break-up.

The initial Moho is a compositional boundary at the base of a primitive continental crust. However, during geologic evolution, other processes such as metamorphism may also affect the lower crust and uppermost mantle to form a Moho, which is not necessarily a boundary between different compositions, but instead a boundary between rocks in different metamorphic state with different seismic and density properties (Mengel and Kern, 1992). Because of the density contrast across the Moho, mafic magma rising from the mantle may experience neutral buoyancy around the Moho, in which case it may pond at this level to create a classic underplate of new crustal material at the pre-existing Moho. Given that the Archaean lithosphere is believed to have been formed by accretion of arc lithosphere and oceanic plateaus (Lee, 2006), underplating may have been a key process in the growth of continental crust in the Archaean because the underplated material may result in secondary melting of parts of the original crust (Fyfe, 1978, 1992). Underplating may also have been a major process in association with the formation of continental flood basalts (Cox, 1980), although its importance has not been confirmed by geophysical imaging so far. The thermal effects of crustal underplating and their consequences for seismic parameters have been discussed in depth by Furlong and Fountain (1986).

Early geophysical tests of the presence of underplated material below the continental Moho were based on acquisition of seismic and gravity data with relatively low resolution, and the results generally were in agreement with models of large, continuous layers of underplated material (e.g. Fowler et al., 1989). Recent seismic experiments at higher resolution have resulted in significantly improved images of the structure of underplated material and mafic intrusions in the continental crust, which has advanced the general understanding of the processes involved. In the following we introduce some of the early results, followed by a presentation and discussion of new findings mainly based on seismic models. They show that underplating is a complicated process which may take many expressions and may not just be related to ponding of mafic magma beneath the Moho. We therefore discuss structure and processes related to a wide variety of structural settings where magmatic processes have altered the depth interval around the Moho in the crust and uppermost mantle.

Identification of underplated intrusive mafic material on the basis of geophysical observations cannot be unique. The discriminators are high P- and S-wave velocity, high Vp/Vs ratio or Poisson's ratio and

high density. However, these characteristics may also apply to granulites from the lower crust which have been metamorphosed into eclogite facies, and to some degree to serpentinized mantle peridotite which nevertheless tends to have lower density than the other rock types. Interpretations therefore have to incorporate other knowledge of the general tectonic setting. Reflection seismic images may further help the identification in cases when the magmatism creates sill like features, which are readily identified by this method, or in cases of large intrusive bodies that have cooled for long time to create a body with smoothly varying properties, which may be reflection free at seismic frequencies. Reflection seismic profiles often image changes from lower crustal reflectivity to a reflection free part of the lower crust, which potentially may indicate the presence of underplated material.

We include a wide range of processes into our definition of underplating, which is "addition of mafic magma to the lower crust and uppermost mantle around the Moho". The original concept of underplating consisted of a simplistic model where magma was ponding just below the Moho. Such underplated layers originally were conceived as having large lateral extent, but this has never been identified by geophysical imaging and may not exist. In the following we argue that underplating in our broad definition takes place in a wide range of tectonic settings (Fig. 1), and it plays a major role in the tectonomagmatic evolution of the lithosphere. A main conclusion is therefore that it is impossible to provide a simple definition of the term underplating.

2. Underplating and lower crustal reflectivity

There is some uncertainty about the origin of the ideas of underplating and when the concept was first proposed. Fyfe (1978) proposes that massive mixing processes occur near the base of the continental crust when mantle magma ponds near the Moho. He finds that such ponded material may be responsible for the formation of granitic magmas by mixing parts of the original magma with remolten crustal rocks, thereby leading to the distinct geochemistry of granites. Fyfe assumes that hotspots are the most likely source of the initial magma. He carries the arguments further (Fyfe, 1992) based on measurements of densities of various magma types from laboratory experiments. He finds that the densities of mantle derived magmas are higher than the average density of the continental crust and close to the density of the lower crust, at least at the advanced stages of magmatism, at the expected temperatures and pressures at the Moho. This provides a strong argument that magmas may accumulate (pond) at the Moho level.

Basaltic magmas may also penetrate into the crust. The style of interaction of high-density, mafic-ultramafic magmas with the crust and the geometry of intrusions which intrude into low-density crustal rocks is controlled by lithosphere rheology (Gerya and Burg, 2007). Elastoplastic rheology that dominates at low temperatures favours upward magma propagation by crustal faulting and results in formation of sills and finger-shaped dykes in the crust. In case of high lithospheric temperatures, magmatic intrusions usually form large flattened mushroom-shaped plutons. Crustal heating caused by underplating and mafic intrusions causes crustal melting and granitic magmatism. It is generally believed that the time-scales for the latter are less than 100,000 years irrespective of tectonic setting, and even when large volume of felsic magmas are produced (Petford et al., 2000). In the latter case, these magmas may solidify as granitic crust (e.g. Coldwell et al., 2011; Douce, 1999; He et al., 2011; Huppert and Sparks, 1988).

There are abundant geologic and petrologic arguments that substantial underplating must exist at the base of the crust (Peltonen et al., 2006; Zheng et al., 2012) but direct geophysical observations in cratonic areas generally indicate that underplated structures are relatively local (Gorman et al., 2002; Korsman et al., 1999; Thybo et al., 2003). However,

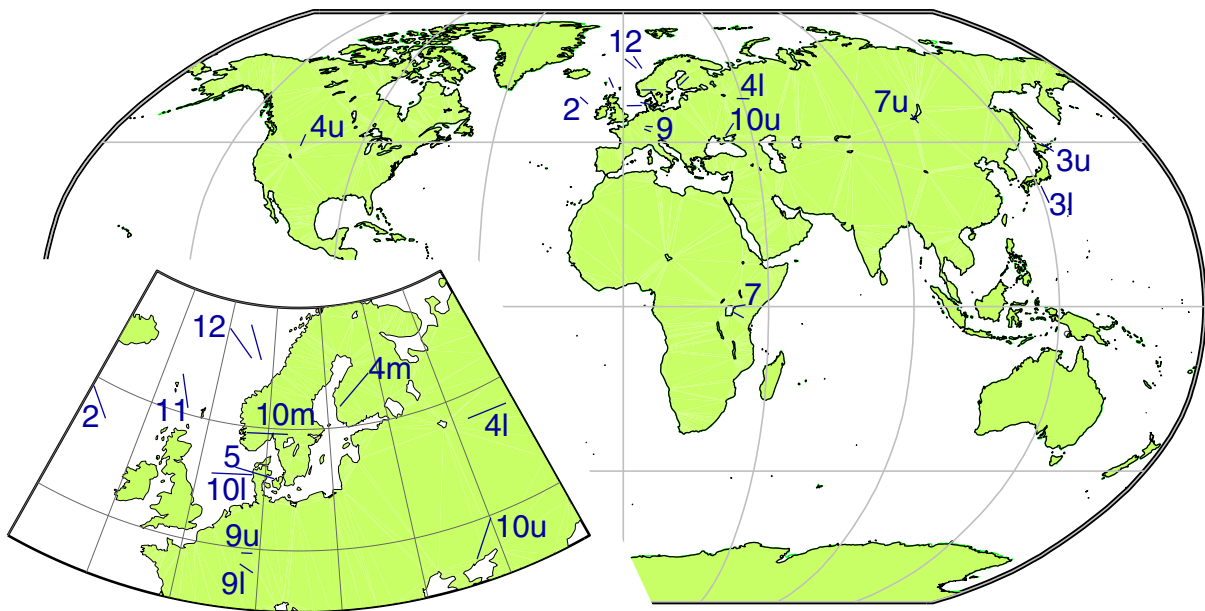


Fig. 1. Map showing locations of seismic profiles discussed in the text and illustrated in Figs. 2–12; inserted map shows zoom to Europe. Profile numbering corresponds to figure number (2–12) and labels correspond to: u – upper, m – middle and l – lower parts of figures.

the extensive occurrence of lower crust with high velocity in cratonic areas (3-layered cratonic crust, e.g. [Garetsky et al., 1999](#); [Meissner, 1986](#)) may also be regarded as indication that it was formed by substantial underplating. If correct, underplating has played a major role throughout the evolution of the cratonic crust. This lower crustal layer is often highly reflective at high frequencies in seismic normal-incidence and wide-angle reflection studies (e.g. [BABEL Working Group, 1993](#); [Cook et al., 1997](#); [Rudnick and Fountain, 1995](#)) and can even be identified in multiply reflected seismic mantle phases ([Morozov and Smithson, 2000](#); [Nielsen and Thybo, 2003](#); [Nielsen et al., 2003](#)).

An early Canadian experiment in the Arctic identified a high velocity lower crust ($V_p = 7.1\text{--}7.7$ km/s) at the Alpha Ridge ([Forsyth et al., 1986](#)). These authors compare this finding to the velocity structure on Iceland and conclude on this basis that substantial magmatism has occurred, and that the region around the Alpha Ridge could have been passing over the presumed mantle plume which is now situated below Iceland. In the light of new knowledge, the assumption of a passing mantle plume may not be required for the production of the magma, because high-velocity lower magmatic crust appears to be a normal feature along rifted continental margins rather than at isolated parts of the margins ([White et al., 2008](#)).

The North Atlantic margins are the classic places where significant underplating related to rifting and break-up processes has been observed. Ray tracing modelling of a profile ([Fig. 2](#)) across the margin at Hatton Bank identifies an up-to 14 km thick zone of unusually high seismic P-wave velocity above 7.3 km/s ([Fowler et al., 1989](#)) and these velocities are further confirmed by seismic data from five expanding spread profiles ([Spence et al., 1989](#)). The occurrence of seaward-dipping reflectors ([Mutter et al., 1982](#)) above the high velocity lower crust indicates a generic connection between the two features, where the original magma fractionated into the main part that today forms the anomalous lower crust and the secondary part that extruded around sea level and now, after cooling and solidification, forms the thin layers of seaward dipping reflectors. The thin seaward dipping reflectors close to the surface are identified by reverberative reflections and their presence causes even reflections from distinct sharp interfaces in deeper levels to appear reverberative. This may be the reason that the reverberative seismic reflections from the underplated layer were not modelled by a layered sequence from the depth region around the Moho. The underplated zone at the North Atlantic margin was therefore originally conceived as being transparent ([Fowler et al., 1989](#)).

An early observation of an underplated layer beneath a sedimentary basin was made at the Valencia Trough in the western Mediterranean Sea. [Collier et al. \(1994\)](#) find indications for relatively thick (up-to 8 km) underplate in the form of a series of subhorizontal reflectors on a length scale of 3–5 km. This underplate in the Valencia Trough mainly exists below the weakly stretched part of the continental crust, whereas the underplated layer is significantly thinner below the areas which have been moderately stretched, and it is absent at a weakly reflective Moho at high stretching factors. This tectonic sequence of variable stretching may be related to post-underplate stretching of the thinnest crust.

Strongly reflective lower continental crust has been widely observed by the extensive seismic reflection imaging that was carried out in the 80–90's by several national and international consortia, both on- and offshore (e.g. [Klemperer and Hobbs, 1992](#); [Meissner and Bortfeld, 1990](#)). The reflectivity was mainly observed in young tectonic provinces, and often in terranes where the last tectonic event had been extensional, e.g. following orogenesis. The interpretations were subject to vigorous debate in the late 80's into the 90's. The reflective lower crust could be explained by igneous,

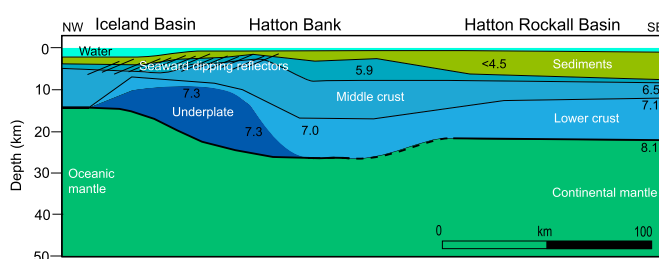


Fig. 2. Seismic P-wave velocity structure of the crust along a profile across the continent–ocean transition at Hatton Bank (after [Fowler et al., 1989](#)). This profile is an early demonstration of underplating at volcanic rifted continental margins. The underplated layer is interpreted from its high seismic velocity and its location below a series of seaward dipping reflectors, believed to have formed during break-up. The underplated layer is inferred as seismically transparent in these early interpretations. Compare with [Fig. 11](#).

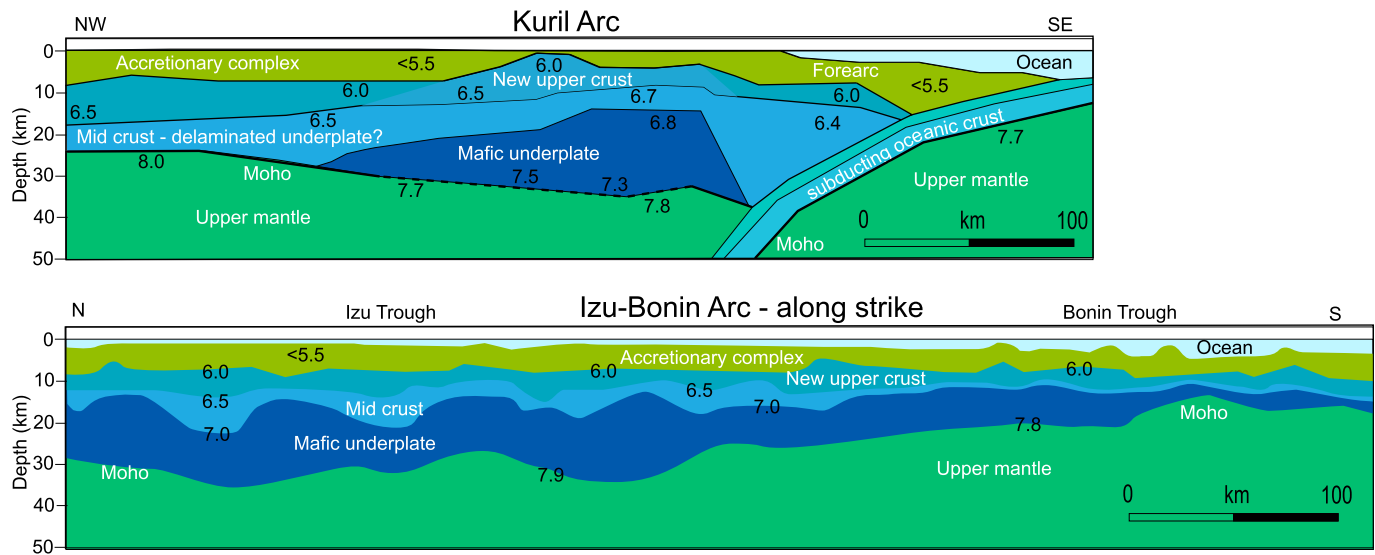


Fig. 3. Upper: Seismic P-wave velocity structure of the crust along a profile across the Kuril Arc in northern Japan (after Nakanishi et al., 2009). The presence of a large body of rocks with high seismic velocity in the lower crust is interpreted as an underplated body which, through fractionation, has provided magma that today forms the new upper crust. This separation between an upper body with “normal continental crust” and a lower body with high velocity is suggested to be an effective factory for producing continental crust (Tatsumi, 2005; Taylor, 1967), although it is unclear if the associated subduction may delaminate the underplated material into the mantle. Lower: Along-strike velocity model of the Izu–Bonin Arc (after Kodaira et al., 2007b), demonstrating substantial variation in crustal structure at two scale lengths: The Izu Arc has relatively thick crust compared to the Bonin Arc, although both arc segments have overall similar crustal structure despite a depth scale factor, and there is significant short-wavelength variation in crustal thickness and structure in both parts of the arc, which has been related to the location of individual volcanoes along the arc.

metamorphic, structural and fluid related models, where the main proposed physical mechanisms were structurally induced anisotropy, metamorphic layering, free fluids, igneous layering, and the juxtaposition of pre-existing heterogeneities by shear zones (Warner, 1990). This author suggests that the latter two mechanisms are the most realistic as the primary causes for the lower crustal reflectivity. However, based on modelling with synthetic seismic sections, Sandmeier and Wenzel (1990) argue that layered variation in quartz content in the lower crust may better explain the observed wide-angle strong P-wave and weak S-wave reflectivity below the Black Forest (the Rhine rift region), whereas Thybo et al. (1994) argue that high fluid content may be the primary explanation for similar relatively weak S-wave reflectivity in the Sorgenfrei-Tornquist Zone (the Baltic Sea region). Meissner and Kuszniir (1987) suggest that often weak rheology is favourable for developing a reflective lower crust. Holliger and Levander (1994) demonstrate by calculation of synthetic seismograms that lithological variation, as observed at the surface in the Ivrea Zone (the Alps), may explain the observed normal incidence reflectivity in the lower crust. Deemer and Hurich (1994) model strong reflectivity with models based on in-situ observations of layered sequences of sills. They conclude that magmatic underplating involves creation of layered structure, and that underplating should not be invoked as an explanation for non-reflective crust except in cases where eclogitization has occurred.

3. Compressional settings

Collision and subduction tectonics is found in a number of plate boundary configurations, including ocean–ocean, ocean–continent and continent–continent transitions. These environments include both extensional and compressional zones each with their tectonic characteristics. Subduction processes are key to the differentiation of the lithosphere into oceanic and continental domains, where the processes of chemical differentiation act as a filter that effectively separates andesitic (granitic, light, upper crustal) from anorthositic (granolithic, heavy, lower crustal) components (Tatsumi, 2005).

3.1. Arc magmatism and crustal formation

The continental crust is believed to originate and grow by addition of andesitic magma (Rudnick, 1995; Taylor, 1967). The magma may originate directly from the mantle by remelting of subducting slabs and the depleted mantle wedge by ascending melt (Kelemen, 1995; Rapp et al., 1999). Although this process cannot be responsible for the formation of the earliest crust, because there were no slabs to melt and no subduction, it may have been the primary process for Archaean crustal production (Taylor and McLennan, 1995). Alternatively, the continental crust may originate from addition of basaltic magma to the pre-existing crust. The resulting magma fractionates into andesitic magma that rises to upper crustal levels and anorthositic magma that remains in the lower crust as an underplate (Taylor, 1967). As the continental crust has lower average density than the oceanic crust, a mechanism is required to separate and remove the lower level rocks of the fractionated magma to form the characteristic low-density continental crust, e.g. by recycling the anorthositic lower crust back into the mantle (Tatsumi, 2005).

In this model, initial continental crust is formed in island arcs (Fig. 3), such as the Izu Bonin and Aleutians (Holbrook et al., 1999; Suyehiro et al., 1996; Taira et al., 1998). The required recycling may be caused by delamination processes. Structural interpretation of seismic profiles at oceanic arcs indicates tectonic separation by obduction of the upper layer and subduction of the lower layer in a crocodile tectonic structure (Lizarralde et al., 2002; Nakanishi et al., 2009); although in more mature continental crust other delamination processes may be important (Artemieva and Meissner, 2012; Kay and Kay, 1993; Mengel and Kern, 1992).

Recent seismic experiments, in particular in subduction systems, show evidence for strong lateral variation in crustal structure along the strike of oceanic arcs (Fig. 3). The crustal thickness along the Izu–Bonin Arc varies between 10–15 km beneath the Bonin Trough and 20–30 km beneath the Izu Trough (Kodaira et al., 2007a; Tatsumi, 2005). The velocity structure in the two parts of the arc is remarkably similar, and the structure in the Bonin Trough appears to be a down-scaled version of the structure of the Izu Trough corresponding

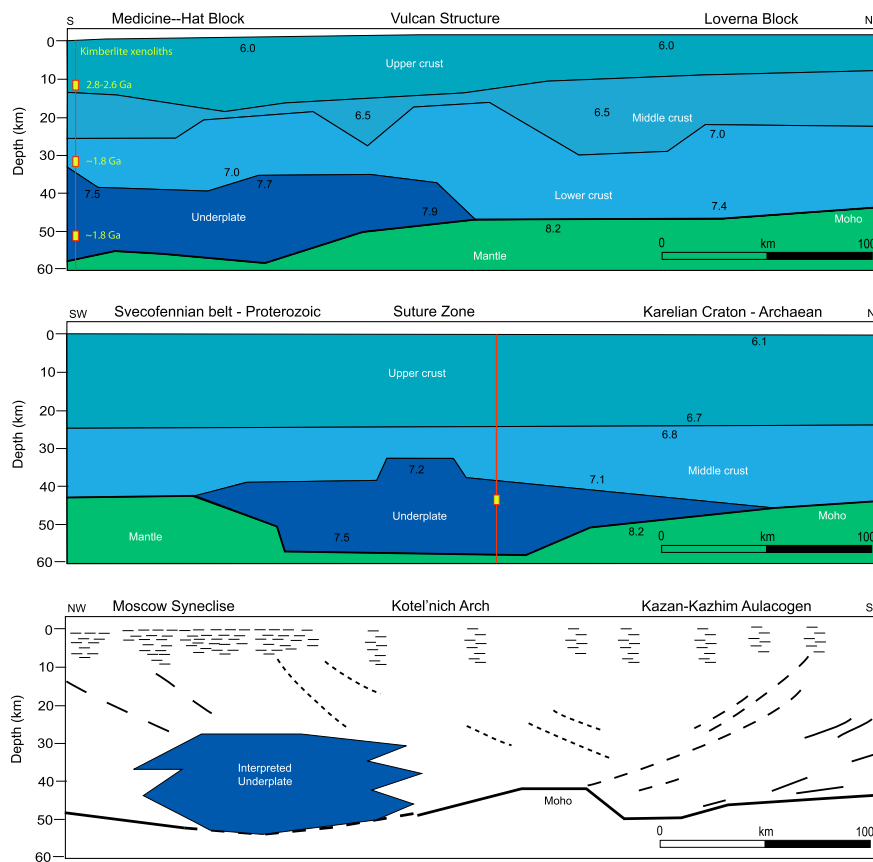


Fig. 4. Seismic velocity structure along crustal profiles in Precambrian terranes including the interpreted underplated bodies. Upper: NS striking profiles across the Canadian–US border between three Archæan crustal domains: the Hearne province (Loverna block) in central Alberta, the Medicine Hat block (MHB) in southern Alberta, and the Archæan Wyoming craton in central Montana (after Clowes et al., 2002). The interpreted underplated layer extends southward from the suture beneath the Phanerozoic-modified Archæan Wyoming craton. Middle: SW–NE striking profile from the Baltic shield, showing an interpreted underplated body in the suture zone between the Archæan Karelian Craton and the Proterozoic (Svecofennian, ca. 1.9 Ga) mobile belt (after Korsman et al., 1999). Lower: Part of the NW–SE striking Tatseis–2003 seismic reflection profile across the Archæan Volga–Uralia craton between the Moscow Basin and the Southern Urals. A reflection-free zone in the lower crust is explained as an underplated body at the transition from a basement high to a 3–4 km deep sedimentary basin (after Trofimov, 2006).

to the change in crustal thickness. It indicates the presence of a well sorted crust, ranging from felsic, over intermediate, to ultra-mafic composition. As the arc is formed by magmatism, the lower ultra-mafic layer can be regarded as an underplate. The seismic data show generally weak reflections from the top and bottom of this layer, indicating that the transitions between the layers are gradual. A remarkable feature of the model is the short wavelength variability in structure (Fig. 3), where zones with low average crustal velocity correspond to the locations of active volcanoes. Such low velocity zones are caused by thickening of the middle crust beneath basaltic arc volcanoes by material with seismic velocities similar to average crustal velocities in continental crust. This observation may indicate that the volcanism launches an effective filtering between felsic, continental, and mafic crustal material. However, overall the profile shows that the bulk composition of the Izu–Bonin arc crust is more mafic than the typical continental crust, such that an efficient recycling mechanism is required to filter the dense part of the arc material from the light material in order to form new continental type crust, if arcs are primary contributors to continental crust formation.

Seismic models perpendicular to island arcs show similar velocity structure in the arc itself with velocities ranging from ca. 2 to 7.5 km/s, whereas velocities are smaller than 6.7 km/s in the backarc region of the Kuril Arc (Fig. 3), and up to 6.9 km/s in the Aleutian arc. In the Aleutian arc, there is some evidence from ringing reflectivity, that the underplating may be taking place in the form of sill intrusions into the lower crust at the back-arc side of the system (Lizarralde et al., 2002).

However, the available seismic profiles from various island arc systems do not provide direct evidence for the processes that may delaminate the underplated mafic to ultra-mafic layer. As such it is an outstanding question if island arc processes may be dominant in building continental crust.

3.2. Underplating in Precambrian crust

Seismic studies in stable continental regions have demonstrated that magmatic underplating is often associated with Archæan crustal blocks that were reworked in the Proterozoic, probably in collisional environments. Xenolith suites from the present lower crust of Precambrian cratons provide invaluable complementary information on the composition, age and the nature of crustal reworking.

One of the well-documented examples of underplating comes from the Lithoprobe seismic refraction profile SAREX across three Archæan crustal domains (Fig. 4): the Hearne province (Loverna block) in central Alberta, the Medicine Hat block (MHB) in southern Alberta and the Phanerozoic-modified Archæan Wyoming craton in central Montana (Clowes et al., 2002). Two significant northward dipping reflectors in the subcrustal lithosphere associated with sutures between the three Archæan blocks are interpreted as relic subduction zones (Gorman et al., 2002). Crustal thickness along the profile increases from 40 km in the north to 50–60 km in the south beneath the MHB and the Wyoming blocks. Domains with the thickest crust

have a prominent 10–25 km thick lower crustal layer with unusually high velocities (Fig. 4) which extends 600 km southwards from the USA–Canada border. The highest velocities, 7.5–7.9 km/s, correspond to the MHB, whereas beneath the Wyoming craton they are 7.0–7.7 km/s (Gorman et al., 2002). In contrast, lower crustal velocities in the northern part of the profile are normal, 7.0–7.4 km/s. The sharp change in crustal thickness and in the velocity structure of the lower crust corresponds to the Vulcan structure, which is a major tectonic boundary between the Hearne and Wyoming provinces clearly defined in potential-field data and interpreted as a zone of palaeocollision. Despite significant differences in the crustal structure, upper mantle velocities are remarkably similar, 8.2 km/s, along the profile.

At depths greater than 35 km and crustal temperatures typical of shield areas with low heat flow, a mixture of mafic granulite (60% at 7.4 km/s) with a significant fraction of ultramafics (40% at 8.2 km/s) can explain remarkably high seismic velocities in the lower crust beneath the MHB, whereas slightly lower velocities beneath the Wyoming block can be explained by the presence of primarily mafic granulites (Clowes et al., 2002). Petrologic and geochronological U–Pb zircon and Nd mineral isochron data for crustal xenoliths from the Montana alkalic igneous province in the MHB indicate that Archaean ages (ca. 2.84–2.65 Ga) are restricted to the upper crustal rocks, whereas lower crustal mafic granulites record Palaeoproterozoic ages (1.85 to 1.72 Ga; Davis et al., 1995; Fig. 4). Based on seismic refraction and xenolith data, the high-velocity lower crustal layer is interpreted to be the result of Palaeoproterozoic magmatic underplating (Clowes et al., 2002).

A somewhat similar crustal structure is observed in the Baltic shield (Fig. 4). There, a deep crustal root (down to ca. 60 km) is associated with the suture zone that separates the Archaean Karelian craton from the Proterozoic (Svecofennian, ca. 1.9 Ga) mobile belt. Crustal thickening is associated with the presence of the 0–25 km thick high-velocity layer (7.0–7.5 km/s) at the base of the crust (Korsman et al., 1999). Gravity modelling and interpretations of seismic data in combination with laboratory measurements of seismic velocities for different lithologies suggest that the velocity profiles cannot be explained by a single rock type, but require the presence of a mixture of mafic garnet granulites, hornblendites, pyroxenites, tonalitic gneiss and some eclogites (Kuusisto et al., 2006). A xenolith suite derived from 40 to 58 km depth indicates that the geophysically determined high-velocity lower crustal layer consists of both Archaean and Proterozoic mafic granulites, whereas the upper crust has zircon ages of up to ca. 3.5 Ga (Peltonen et al., 2006). Thus, the presence of the high-velocity lower crustal layer is attributed to Proterozoic basaltic magma underplating and mixing with pre-existing Archaean mafic granulites as a response to accretion of the Svecofennian arc complex to the craton margin.

In the absence of xenolith data from the inner parts of the East European Platform, geodynamic interpretations of seismic data are less constrained. The ca. 1000 km long Tatseis-2003 seismic reflection profile from the Moscow basin to the Southern Urals across the Archaean Volga–Uralia craton shows a highly heterogeneous crustal structure with a sharp change of reflectivity at Moho, at depths varying from ca. 40 to 50–55 km (Trofimov, 2006). Transparent lower crust in the NW part of the profile may be explained by the presence of underplated material at the transition from a basement high to a 3–4 km deep sedimentary basin (Fig. 4).

It is noteworthy that crustal underplating or the presence of high-velocity lower crust is not unequivocally documented for modern collisional orogens. For example, the velocity structure of the Altiplano crust, where crustal thickness is 60–65 km, is more consistent with felsic composition than the presence of underplated mantle derived material or magmatic intrusions from the mantle (Swenson et al., 2000).

4. Extensional settings

Extension of continental lithosphere leads to the formation of rift zones, wide extensional areas such as the Basin and Range Province and continental shelves, and ultimately to new oceanic crust (Ruppel, 1995). Extensional tectonics is to a variable degree accompanied by magmatic activity, depending on the stretching amount, the amount of fluid in the system, and the amount of heat being transferred from below to the extended area. Early models of continental extension considered two end members in terms of active or passive rifting (Sengor and Burke, 1978). The main forces behind passive rifting originate from tectonic processes at distant plate-boundaries causing deformation within the plate interior (Artemjev and Artyushkov, 1971). In contrast, active rifting is initiated by the stretching caused by thermal uplift due to heating of the lithosphere from below. Both models lead to the formation of rift grabens which are elongated depressions in the Earth's surface, and with time become filled with sedimentary and volcanic material, as is presently observed at e.g. the Baikal, East African, Rhine Graben and Rio Grande Rift Zones. If the tectonic situation is favourable for continued extension, new oceanic lithosphere may be formed as currently observed in the Red Sea. Recent research indicates that rift evolution usually involves both mechanisms, where one of the two end-member mechanisms may be dominant at the early stages of rifting.

Regions of extended crust and lithosphere will usually be subject to magmatism, but significant variation in the intensity of volcanism is observed between rift zones, ranging from dry (little or almost no volcanic activity) to wet (intensive volcanic activity) rifting. Overall seen, there may be a tendency that the magmatic activity begins earlier in the case of active rifting, due to early heating from below, than in the case of passive rifting where mantle melting is mainly caused by decompression from the lithospheric thinning. However, recent studies have indicated that the melting in extended regions may be caused by simultaneous decompression and heating in the melting zone of the mantle (Thybo and Nielsen, 2009). The structures caused by magmatic activity assume many forms, including surface volcanoes, dikes, sills, batholiths and perhaps massive underplating, e.g. at the crust–mantle boundary. Models of the shape of such intruded structures have developed substantially recently, as the resolution of seismic imaging has improved with time.

4.1. Wide extensional areas

The large Basin and Range Province has undergone substantial stretching and the area is underlain by upper mantle with P-wave velocity less than 8.0 km/s. Estimates of extension based on various tectonic models provide values of up to several hundred percent, which corresponds to a beta factor locally higher than 3 (Parsons et al., 1996; Wernicke, 1985). One may expect substantial decompressional melting in such an environment, but all seismic profiles indicate that any possible underplated layer cannot exceed a thickness of 4 km (McCarthy and Parsons, 1994). This small amount of magma supplied may suggest that the stretching has been passive or that, possibly, a previously existing underplated layer has been delaminated. Even though the available seismic data has been scrutinised for evidence of high-velocity and reflective lower crust, there is only sporadic seismic evidence for a ~2 km thick layer that may be regarded as a mafic underplate (Suetnova et al., 1993). These authors find that this thin layer is highly reflective and interpret the reflectivity by a series of thin mafic sills at the base of the crust. A similar conclusion has been made from an analysis of seismic P- and S-waves from the Basin and Range (Goodwin and McCarthy, 1990). These conclusions are supported by xenolith studies that indicate an up-to 2 km thick layered sequence at the Moho with interlayered crustal and mantle derived material (McGuire, 1994).

4.2. Large batholiths and sills in a Moho transition zone

The Bushveld intrusion in the cratonic lithosphere of southern Africa represents a massive intrusion of large continuous volumes of magma in the crust. The intrusion represents the root of a presumed continuous magma chamber which contains equal amounts of felsic and mafic material, probably formed by fractionation of magma and remelting of the surrounding continental crust during the long cooling period following intrusion at ca. 2.06 Ga (Scoates and Friedman, 2008). Its present surface expression is in two large complexes, separated by about 250 km, and it is debated if they represent one continuous large igneous intrusion or if they are separate structures. Recent gravity interpretation, with constraints from receiver function observations of crustal thickness, indicates that it may be one continuous structure with a volume of more than 1 mln km³ (Cawthorn and Webb, 2001; Webb et al., 2011), but new receiver function calculations indicate that the two complexes are independent (Youssof et al., this volume). Although the Bushveld intrusion may not have solidified at Moho level, this large igneous intrusion has undoubtedly served as inspiration for models of continuous igneous intrusion around Moho at e.g. continental margins.

Another example is the gabbro–anorthosite–rapakivi Korosten pluton in the Palaeoproterozoic block of the Ukrainian shield. There a ca. 10 km thick high-velocity layer (7.6 km/s) is observed above the Moho in an area with local Moho deepening (Thybo et al., 2003). Its presence and the internal stratification in the layer may indicate intrusion of mantle-derived melts into the lower crust and around the original Moho as well as lower crustal melting during the emplacement of the pluton (ca. 1.75 Ga) or during the formation of the adjacent Devonian–Triassic Pripyat Trough.

An extremely large, continuous mafic intrusion in the crust has been observed by seismic data in the Norwegian–Danish Basin at the edge of the Fennoscandian shield (Fig. 5). It is interpreted as a large batholith (extremely thick underplate; Thybo and Schönharting, 1991), which has been observed along three seismic profiles and its size has been estimated by interpolation constrained by gravity data. Its dimensions are 20 km thick, 20–30 km wide, and more than 110 km long with a total volume of at least 60,000 km³ (Sandrin and Thybo, 2008a, b). The structure has very high seismic velocity – ca 6.8 km/s at a depth of 10 km and between 7.0 and 7.7 km/s at Moho level, with highest Vp in the central part of the structure. The corresponding velocities in the surrounding crust are 6.3 km/s and 6.8 km/s, demonstrating a large velocity contrast across the edges of the structure. It is remarkable that the interpreted batholith is seismically transparent for both normal-incidence and wide-angle seismic reflections in a wide frequency range. The seismic transparency indicates that it has been cooling as

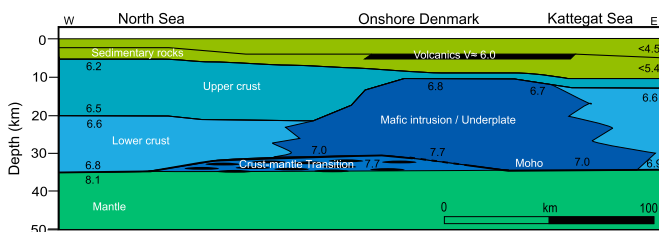


Fig. 5. Seismic P-wave velocity structure of the crust along a profile across a large positive gravity anomaly in the Norwegian–Danish Basin (after Thybo and Nielsen, 2012). The large high-velocity body is interpreted as a mafic intrusion (exceptionally thick and large underplate). The presence of volcanic rocks in the late Palaeozoic sedimentary sequence suggests a similar age of the intrusion. The Moho cannot be detected seismically at the zone with the highest velocity of 7.7 km/s (see Fig. 6), probably due to a gradual transition between crust and mantle at this interpreted feeder channel. The Moho is a layered, highly reflective transition zone over more than 120 km towards the west from the intrusion (see Fig. 6). This zone is interpreted as a series of sill-like intrusions that may all have originated from the large intrusive body.

one large magma chamber over a long period to allow gradual fractionation between melt and solidified magma. Using the method proposed by MacLennan and Lovell (2002) the cooling time is estimated to be up-to 10–20 My. Reflections from the Moho are observed along the strike of the structure (Fig. 6), except in the central part where the seismic velocity (7.7 km/s) is comparable to the unusually low velocity (7.8 km/s) of the uppermost mantle (Fig. 6). This observation has been interpreted as evidence that the magma intruded into the crust along feeder dikes in the middle of the structure (Thybo and Nielsen, 2012). Other similar, coeval batholiths were identified in the Norwegian–Danish Basin (Heeremans et al., 2004; Thybo, 2000; Timmerman et al., 2009). Lithosphere heating by such densely distributed, massive intrusions during the late Permian must have caused substantial surface uplift followed by erosion. Magma solidification and cooling may have led to the initiation of the wide, regional subsidence of the Danish Basin during the Triassic (Sandrin and Thybo, 2008b).

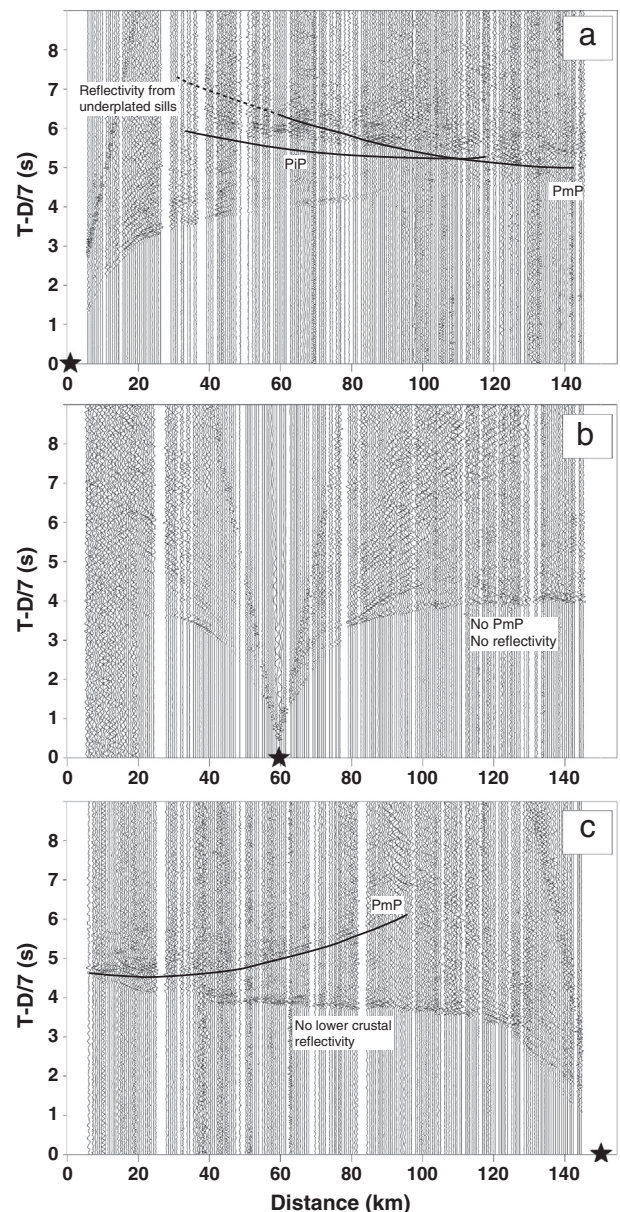


Fig. 6. Seismic sections along the profile in Fig. 5 (after Sandrin et al., 2009). a. Strong reflectivity from the layered zone around Moho. b. Lack of PmP-reflection from the Moho at the zone with the highest velocity in the large intrusive body. c. Short, non-reverberative PmP-reflection from Moho in most of the large intrusive body.

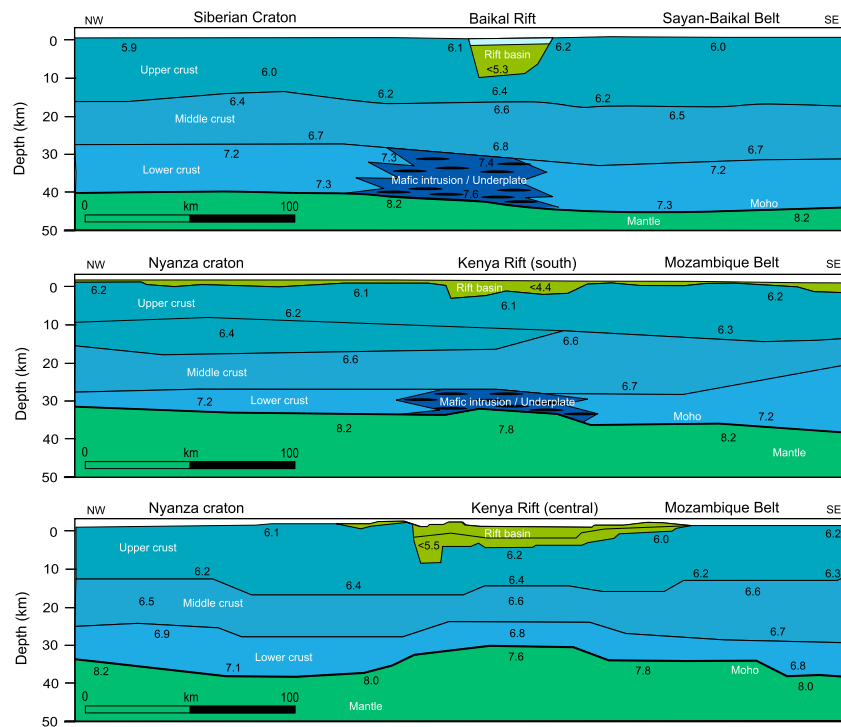


Fig. 7. Seismic P-wave velocity structure of the crust along three profiles across presently active rift zones: The Baikal Rift zone in Siberia (upper, after Thybo and Nielsen, 2009), southern Kenya Rift (centre, after Birt et al., 1997) and the central Kenya Rift (lower, after Maguire et al., 1994). Around the rift grabens, the profile across Baikal Rift shows a large high velocity body in the lower crust, the profile across southern Kenya Rift shows a similar, smaller body above a slight uplift of the Moho, whereas the profile across central Kenya Rift shows larger Moho uplift and no high-velocity lower crustal body. The high velocity bodies are strongly seismically reflective, and are interpreted as series of mafic intrusions in the form of sills that compensates fully or partially for the expected crustal thinning at the rift zones in the first two cases.

Over a ca. 130 km long zone extending westward from the edge of the main batholith in the Danish basin, the Moho transition shows a remarkable signature as a ca. 5 km thick highly reflective zone with a low seismic velocity (Thybo and Nielsen, 2012) (Fig. 5). The reflectivity has been interpreted to be caused by layering corresponding to a sequence of sill-like features that intruded the lowest crust or along the crust–mantle boundary during the late stages of cooling and solidification of the batholith. The layering (Fig. 5) is observed from a high-amplitude, ca. 1 s “ringing” wave-train between the PiP reflection from the top of the reflective zone and the PmP reflection from the base of the zone; the latter may represent the transition into mantle proper (Fig. 6). By modelling with synthetic seismograms, these observations are explained by layering of ca. 500 m thick lower crustal rocks with velocity of 6.7–6.8 km/s and mantle derived mafic rocks with velocity of ca. 7.3 km/s (Sandrin et al., 2009).

4.3. Rift zones: magma compensated crustal thinning

4.3.1. Modern rifts

The presently active rift zones, despite they were initiated at approximately the same time about 40–30 My ago, show significant differences in magmatic activity. Estimates of the surface volcanic output from the four major rift zones vary considerably from some 144,000 km³ of volcanics erupted since the early Miocene in the East African Rift (Williams, 1972) to 5000–6000 km³ in the Baikal Rift system (Kiselev, 1987). Furthermore, the volcanic activity in the East African Rift Zone was very early and began a few million years before the first observed faulting in most parts of the rift (Morley, 1994; Smith, 1994), whereas volcanism apparently began at about the same time as faulting at Baikal Rift Zone (Kiselev, 1987).

At Baikal Rift Zone, there is strong evidence for the presence of underplated material in the form of mafic sills in the lower crust directly below the graben and slightly displaced to the northwestern side (Fig. 7). Strong wide-angle seismic reflectivity is observed from

a 50–80 km wide zone in the lower crust, whereas the lower crust is remarkably non-reflective outside this zone (Fig. 8). The reflectivity can be modelled by fine scale heterogeneity with typical vertical scale lengths of 300–500 m and large reflection coefficients (Nielsen and Thybo, 2009a,b). The lateral extent of the reflective zone can be readily modelled from the observed seismic sections because the reflectivity terminates abruptly with offset (Fig. 8). Outside the reflective zone, the Moho is distinct with a relatively long waveform, indicative of a complex sub-Moho mantle (Fig. 8). The average velocity of the reflective zone (7.4–7.6 km/s) is consistent with the presence of 50% of mafic material intruded into a stretched pre-existing lower crust with properties similar to the adjacent Siberian Craton (Cherepanova et al., this volume). Despite the ongoing rifting, the uppermost mantle shows no sign of decreased velocity (constant across the rift zone at 8.2 km/s down to at least 60 km depth). The Moho shows no sign of uplift below the rift zone, instead its depth varies smoothly between 40 km below the Siberian Craton and 43 km below the Sayan–Baikal fold belt to the SE of the rift zone. The presence of a deep rift graben above a flat Moho may be explained by compensation of crustal thinning by magmatic intrusions (underplating) in the lower crust and around the Moho (Thybo and Nielsen, 2009). The individual reflective bodies may be 300–500 m thick mafic sill-like intrusions which fractionated during solidification. Assuming that the high-velocity underplated layer also has high density, gravity modelling indicates crustal isostatic equilibrium across the rift zone (Thybo and Nielsen, 2009).

Similar observations have been made at the southern part of the Kenya Rift Zone (Fig. 6) where a strongly reflective lower crust is observed directly below the graben (Birt et al., 1997). There is almost no Moho uplift in this area and its amplitude is much smaller than expected from the volume of the subsided rift graben. There seismic reflectivity from the lower crust below the rift graben is qualitatively different from outside of the rift zone. The reflective lower crustal layer is observed above a low-velocity uppermost mantle zone ($V_p < 7.8$ km/s). The P- and S-wave reflectivity from the lower crustal zone is similar and

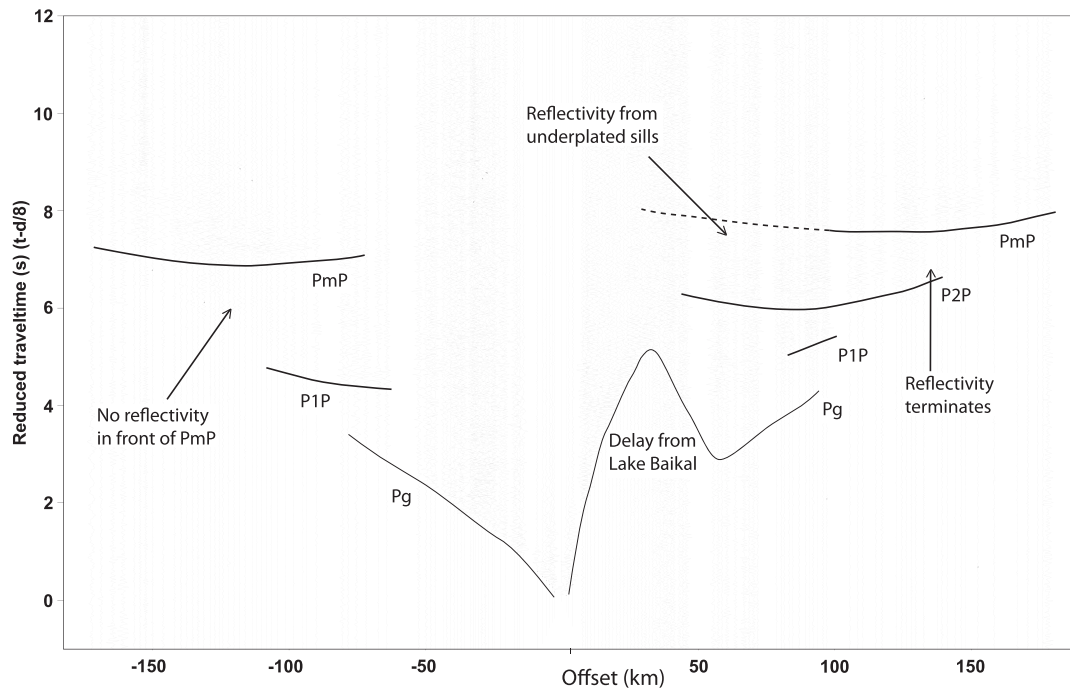


Fig. 8. Seismic section along the profile across Baikal Rift (Fig. 7; after Nielsen and Thybo, 2009a). It illustrates the strong seismic reflectivity from the high-velocity body (HVB) in the lower crust. Its abrupt termination makes it possible to delimit the reflective zone to the HVB. Seismograms show no sign of similar reflectivity for waves that have not been affected by the HVB.

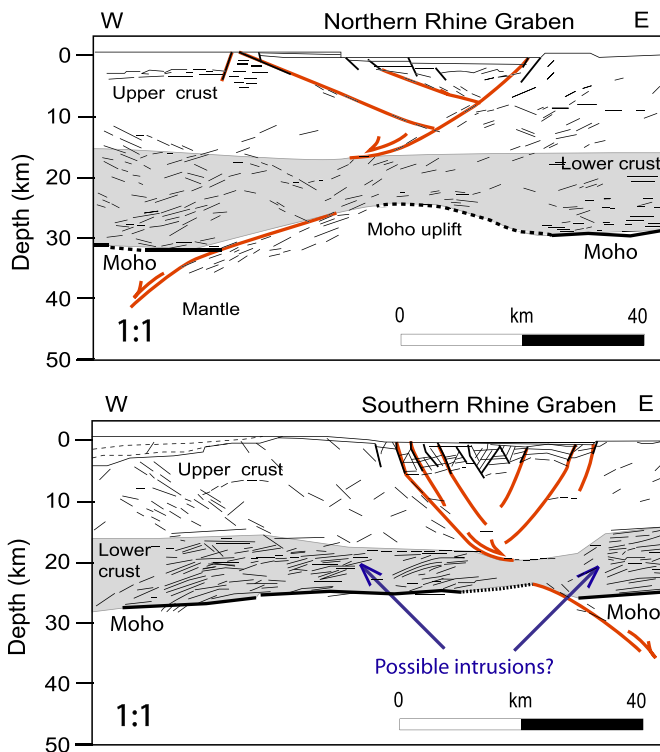


Fig. 9. Line-drawings of crustal seismic normal-incidence reflection sections along profiles across the Rhine Graben (after Brun et al., 1992). The two sections show the presence of a 3–5 km deep graben, an almost non-reflective upper crust and, to a variable degree, a reflective lower crust. The northern profile indicates substantial Moho uplift below the graben, whereas the southern profile shows an almost flat Moho. Lower crustal reflectivity is strongest in the southern profile, which may indicate the presence of underplated mafic sills and suggests that magmatic compensation of the crustal thinning may have been effective here. Interpretation in red is after Brun et al. (1992), whereas the indicated interpreted intrusions (in blue) are after Wenzel and Sandmeier (1992).

indicates very large reflection coefficients (corresponding to velocity contrasts of around 1 km/s with an average P-wave velocity > 7.2 km/s), which is consistent with the presence of highly fractionated mafic sills in the lowermost crust (Thybo et al., 2000). However, the uppermost mantle has low velocity of 7.8 km/s below the rift graben as compared to 8.0 km/s outside the rift zone (KRISP Working Party, 1991). This is different from the Baikal Rift and may reflect the pronounced difference in the amount of magma that has erupted at the two rift zones.

About 250 km north of the southern Kenya profile, another seismic profile indicates a different crustal structure in central Kenya, although the quality of the seismic data is lower than along the southern profile (Figs. 5, 6). It shows a Moho uplift of 4 ± 2 km below the rift graben and no sign of high seismic velocity in the lower crust (Braile et al., 1994; KRISP Working Party, 1991; Maguire et al., 1994). Differences in crustal structure along the Kenya Rift indicate that crustal thinning may in some cases be compensated by magmatic intrusions as along the southern profile and at Baikal rift zone, and that the crust also may remain thinned as along the northern profile.

Crustal structure of the northern Rhine Rift is close to a classic pure shear model (Brun et al., 1992; Fig. 9). The graben is surrounded by rift shoulders some 4–500 m above the graben with peaks up to almost 1000 m above mean sea level. The Moho is uplifted from 30 to 32 km depth by some 6 km below the up-to 3.4 km deep sediment-filled graben. The lower crust on the eastern side of the rift graben is strongly reflective in a normal-incidence reflection seismic section. Similar reflectivity is observed in seismic refraction data with lower frequency content (Sandmeier and Wenzel, 1990).

About 150 km further south, the Rhine Rift graben is only ~2 km deep, perhaps due to a post-subsidence uplift related to a peak in magmatic activity at ~17 Ma (Brun et al., 1992). A normal incidence reflection seismic profile across the southern Rhine Rift shows a lower crust which is apparently seismically transparent below the graben but reflective on both sides (Fig. 9). The initial interpretation assumed that the transparent interval included a Moho uplift similar to the northern Rhine Rift (Brun et al., 1992), but this could not be demonstrated due to the lack of seismic refraction data. We find it likely that the Moho is approximately flat across the southern rift

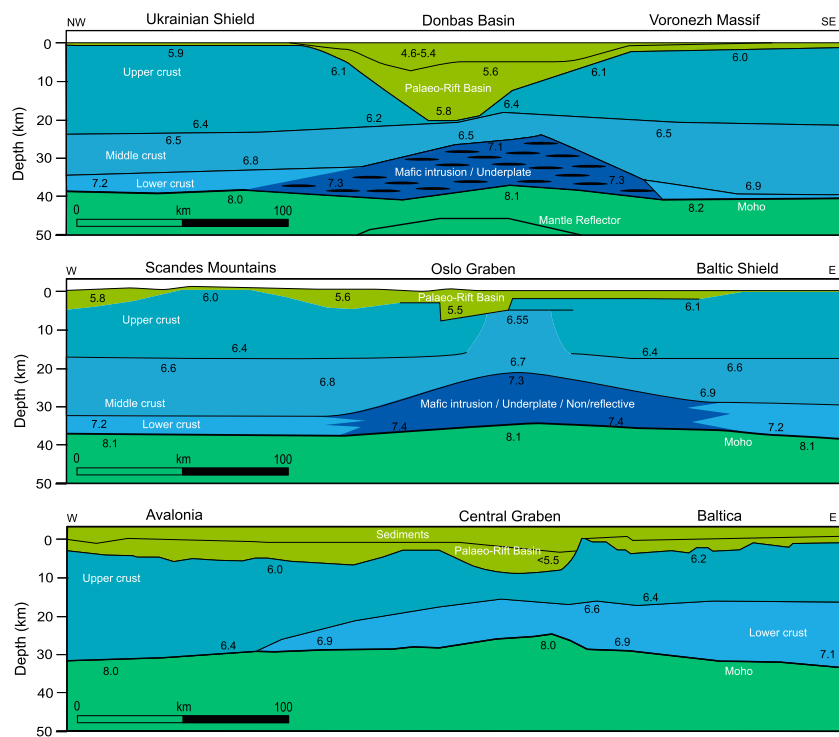


Fig. 10. Seismic P-wave velocity structure of the crust along profiles across the Donbas Basin in Ukraine (upper, after DOBREfraction Working Group, 2003), Oslo Graben in Norway (centre, after Stratford and Thybo, 2011) and the Central Graben in the North Sea (lower, after Nielsen et al., 2000). Avalonia is a former microcontinent that was amalgamated to Laurentia and Baltica during the Caledonian Orogeny. Around the rift grabens, the profiles across the Donbas Basin and the Oslo Rift show large high velocity bodies in the lower crust (indicative of magmatic compensation of crustal thinning), whereas the profile across the Central Graben in the North Sea shows some Moho uplift and no high-velocity lower crustal body, indicative of a classic simple shearing. The interpreted underplated body is highly reflective below the DonBas basin, indicative of 3–500 m thick sills (Lyngsie et al., 2007), and transparent below Oslo Graben, indicative of long term cooling of a large magma chamber.

zone. Given that the crust in the south is generally about 3 km thinner than in the north, stretching mechanisms could also be variable along the rift.

The reflective lower crust on the sides of the southern Rhine Rift may be attributed to abundant mafic sills, similar to the interpretations of the Kenya and Baikal rift zones. In this model, lithosphere stretching and possible heating from below produced mantle melts which have risen to the lower crustal level. Due to the geometry of the extensional faults, the melts may not have migrated to a zone directly below the main graben structure, but instead found it favourable to migrate side-ward. Geochemistry of surface volcanics indicates that some fractionation may have taken place at lower crustal level (Wenzel and Sandmeier, 1992). The magma compensation of the stretching volume has been uneven along the rift due to a variable amount of magma supply. A refraction profile in the Black Forest Mountains on the eastern flank of the rift graben indicates reduced Vp velocity in the lower part of the upper crust and relatively high Vp in the reflective lower crust (Gajewski and Prodehl, 1987). Wide-angle reflectivity indicates a gradual velocity transition instead of a sharp discontinuity at both the top and bottom (Moho) of the reflective lower crustal layer (Sandmeier and Wenzel, 1990). However, the entire lower crust is highly reflective, possibly with a slightly different frequency content of the reflected signals from within the layer and from the Moho transition (Fig. 9). The Moho appears as a non-distinct, gradual transition zone. Stronger reflectivity of the lower crustal layer in P- than in S-wave has been interpreted by high quartz content without explanation of its origin, although the presence of free fluids in the lower crust cannot be ruled out (Sandmeier and Wenzel, 1990). Wenzel and Sandmeier (1992) argue that the reflective lower crust beneath the Black Forest may be caused by mafic intrusions associated with the Rhine Graben rifting, which later were subject to metamorphism. As a result, water was expelled and today resides only in the lower part of the upper crust, where it may be the cause of the observed reduced velocity. This explanation is

attractive, since Moho uplift is not observed in the southern Rhine graben and it is small in the northern section. The latter could be explained by magmatic intrusions from the mantle which compensate the Moho uplift expected from lithosphere stretching. In such a scenario, the lower crustal reflectivity could indicate the presence of mafic sills.

4.3.2. Palaeo-rifts

Palaeo-rift zones show substantial variation in crustal structure. Two end members include the Palaeozoic–Mesozoic Central Graben in the North Sea (Fig. 10) and the 1.1 Ga North American Midcontinent rift. The Central Graben is underlain by a slightly uplifted Moho (Nielsen et al., 2000), and there is only little sign of magmatic intrusion in the crust observed primarily as sub-vertical structures that probably intruded along fault zones (Lyngsie and Thybo, 2007). In contrast, the Midcontinent rift shows strong variability in lower crustal normal-incidence and wide-angle seismic reflectivity, which suggests the presence of large amounts of underplated material (batholithic structures surrounded by sill like features), which intruded during the rifting episodes. Seismic models complemented by gravity constraints indicate that an up-to 15 km thick zone in the lower crust has been affected by underplating processes (Behrendt et al., 1990; Hinze et al., 1992).

Another example of the second type is the Palaeozoic DonBas rift in Ukraine (Fig. 10), where the Moho is almost flat across the rift zone, and a high-velocity, highly reflective lower crustal zone is slightly displaced to the south of the graben (DOBREfraction Working Group, 2003; Maystrenko et al., 2003). The reflectivity and high velocity of the latter zone have been interpreted as evidence for substantial magmatic intrusion during rifting, which has compensated for the crustal thinning (Lyngsie et al., 2007). The presence of intruded sills intermixed with feeder dikes is consistent with the measured seismic anisotropy across the DonBas zone (Meissner et al., 2006).

The Palaeozoic Oslo Graben rift structure (Fig. 10) developed simultaneous with the intrusion of the mafic dikes in the Norwegian–

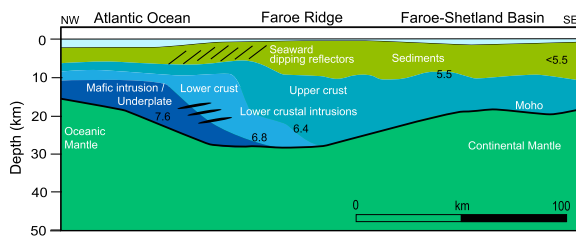


Fig. 11. Seismic P-wave velocity structure of the crust along a profile across the continent–ocean transition at the Faroe Islands (after White et al., 2008). The overall velocity structure is similar to the profile from the Hatton Bank in Fig. 2, but a coincident normal-incidence reflection profile demonstrates that the continent-ward part of the underplated body is highly reflective. The reflectivity is confined to a landward part of the high-velocity lower crust, which indicates that it originates from rifting related intrusions similar to observations at presently active and extinct rift zones (Figs. 5 and 7). The break-up underplate may have been in the form of a massive intrusion, similar to the large crustal intrusion in Fig. 5.

Danish basin, and was probably caused by the same stress field, possibly in association with a proposed hotspot in the area (Neumann et al., 1992, 2004). The rift zone is regarded as a classic example of rifting and the presence of a high density “rift pillow” in the lower crust was interpreted by early gravity studies (Ramberg and Smithson, 1971). However, later studies favour wet mantle melting without control from a hot mantle plume (Pedersen and Vanderveek, 1994) as indicated by the small volume of volcanics and lack of surface uplift before rifting (Heeremans et al., 1996). Yet, until the acquisition of seismic data, it has been impossible to assess the amount of magmatic additions to the crust. Recent seismic data across the southern Oslo Graben show slight crustal thinning (ca. 2 km) below the graben and elevated seismic velocities below a depth of 5 km (Stratford and Thybo, 2011). The data allow interpretation of S-wave velocity, and the observed Poisson’s ratio is consistent with the presence of mafic addition in the high velocity zones of the middle and lower crust. The lower crustal high-velocity zone is wider than the surface expression of the rift graben, whereas the velocity structure indicates a narrow zone of magmatic addition to the middle crust which is located directly below the graben feature.

4.4. Volcanic rifted continental margins

It is generally believed that many volcanic rifted continental margins (also often termed passive margins) have been heavily underplated

during the late stages of rifting and break-up. For example, the presence of a 15–20 km thick high-velocity lower crustal layer is documented for the northern Grenville Province in two Lithoprobe onshore–offshore refraction seismic lines in the Eastern Canadian Shield (ECSOOT, Funck et al., 2001). The lower crustal layer is stratified and has velocities of 7.1–7.4 km/s in the upper part and 7.6–7.8 km/s in the lower part. Based on the correlations of on-shore and off-shore tectonic structures, the high-velocity layer is interpreted as underplating formed during Iapetan rifting and it is proposed that it may be a continuous feature along the Grenvillian passive margin.

Recent studies of the underplated zone of the crust at the North Atlantic margins, by integration of normal-incidence and wide-angle reflection seismic data, have demonstrated that a part of the high velocity zone is seismically reflective, which may suggest that the underplated material intruded as sills into the lower crust above the Moho (White et al., 2008; Fig. 11). This interpretation was substantiated by similar observations at active rift zones, including the Baikal Rift Zone, and the formation of this reflective zone could well have begun during the rifting phase (Thybo and Nielsen, 2009). However, the presented seismic section (Eccles et al., 2011; White et al., 2008; Fig. 11) indicates that the seismic reflections from the lower crust are confined to the landward side of the underplated layer. We speculate that the sill like intrusions in the lower crustal reflectivity zone originate from the time of rifting. The main underplated body is reflection-free and indicates that a substantial volume of magma was supplied at the time of break-up, which must have been followed by a long-lasting solidification of the magmatic body, like for the large crustal intrusion in the Norwegian–Danish Basin (c.f. Fig. 5). Also the presence of seaward dipping reflectors at the margin, interpreted as lava flows during break-up, suggests intrusion of substantial amount of magmas during the break-up. The flow of extrusive material covers a 100 km wide region close to the Faroe Islands whereas the sills in the lower crust spread only over a ~50 km wide zone, despite occupying a much larger volume.

The underplated layer and sill-like intrusions are observed at the continent–ocean transition (COT). Significant underplating at the COT around the North Atlantic was reported in a number of studies (Holbrook et al., 2001; Mjelde et al., 2009; Voss et al., 2009), although the internal reflective character may not have been imaged due to lack of resolution. As such it is unknown if underplated intrusions at the COT always include series of sill like bodies or if they may also be represented by large continuous volumes (large magma chambers) only. The seismic sections from the Vøring margin at the Norwegian shelf (Fig. 12) show the presence of very large, more than 10 km thick,

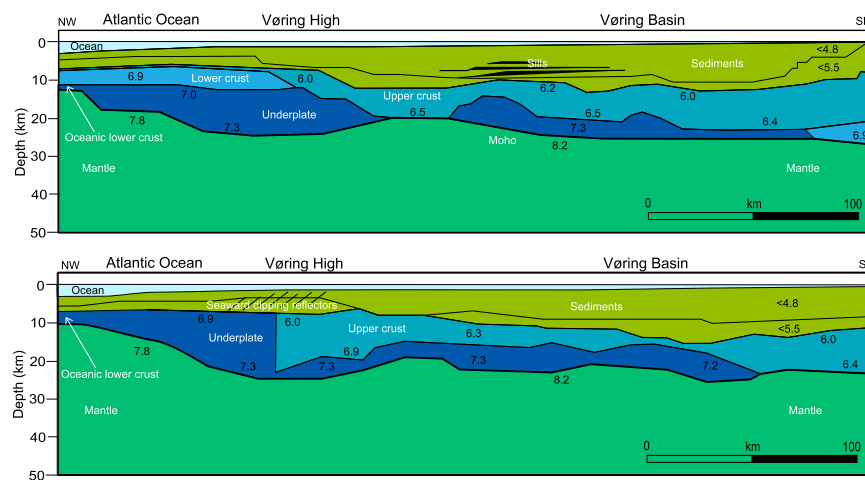


Fig. 12. Seismic P-wave velocity structure of the crust along profiles about 100 km apart (after Mjelde et al., 2009) across the Vøring margin (at the Norwegian shelf) in the North Atlantic, ca. 500 km north of the profile in Fig. 11. This margin shows more substantial underplating at the continent–ocean-transition than in the Faroe profile (Fig. 11). Seaward dipping reflections are observed above the underplated body in one profile. Significant underplating is also inferred from observation of large high-velocity bodies below the Vøring Basin on the Norwegian Shelf.

high-velocity bodies at COT that may mainly consist of underplated material related to the break-up and margin formation. Further landward, thinner high-velocity bodies may represent underplated material related to the formation of the wide, deep Vøring Basin. Comparison of P- and S-wave models shows that the high-velocity zone has relatively high Poisson's ratio corresponding to a V_p/V_s larger than 1.7. As the P-wave velocity is higher than 7.0 km/s, this observation is consistent with a mixture of felsic continental crust and mafic intrusions (Eccles et al., 2011; Mjelde et al., 2003; Raum et al., 2005). An extensive study of P- and S-wave velocity structure along several profiles on the Vøring margin indicates substantial variation in the thickness of the underplated layer, although generally 2–6 km thick, and very high V_p/V_s ratios of around 1.8 with P-wave velocities higher than 7.2 km/s (Mjelde et al., 2003). Recent assessment of seismic data from continental margins globally has indicated that some of the interpreted underplated material, in particular below basins on the continental shelf, may rather be interpreted as lower crustal material that has been metamorphosed into eclogite facies (Mjelde et al., this volume). This interpretation may also apply to the lower crust below the Vøring Basin in the two profiles in Fig. 12.

By comparison of four profiles perpendicular to the coastline of the southern part of eastern Greenland, Holbrook et al. (2001) observe that the volume of the underplated layer depends strongly on the distance from the Greenland–Iceland Ridge, which is believed to consist of more than 25–30 km thick oceanic crust. The thick oceanic crust may be explained by very high melting temperatures in the mantle (Parkin and White, 2008), but the origin of these high temperatures is not well known. Out to ca. 500 km away from the Greenland–Iceland Ridge, the underplated layer is >30 km thick, whereas it is only 18 km thick 500–1100 km further away. This variation may be explained as the effect of increasing distance from the presumed hotspot at Iceland, but the geodynamics of this model lacks documentation (Artemieva and Thybo, 2008). Anomalous crust of the Greenland–Iceland Ridge may extend to the Faroe–Iceland Ridge and possibly also form the Baffin Bay Ridge (Artemieva and Thybo, this volume).

Similarly, underplated structure has been observed at other volcanic rifted margins, e.g. in the Central Atlantic (Kelemen and Holbrook, 1995), South Atlantic (Dupre et al., 2011), parts of the South China Sea where the V_p/V_s ratio is also very high (Zhao et al., 2010), and the polar Alpha Ridge (Funck et al., 2011). However, the coverage by seismic profiles at other margins is sparser than in the North Atlantic Ocean. Seismic data indicate that some volcanic margins may not have been subject to underplating of the stretched continental crust, e.g. at the Labrador Sea (Keen et al., 2012) and the northern South China Sea (Qiu et al., 2001).

5. Large igneous provinces, plumes, and cratons

Magmatic underplating processes are essential for continental flood basalt volcanism. It has been suggested that the major part of the magma that reach the crust may solidify as underplated material and remain hidden, even in continental flood basalt provinces, where large volumes of magma reach the surface (Cox, 1980). Cox (1993) discusses geologic arguments that may support the importance of underplating for continental flood basalt volcanism. These include (i) observation of gabbro fractionation in erupted basaltic sequences, which may provide indication for the amount and mass of concealed material; (ii) observation of large volumes of rhyolite along the continental margin in southern Africa which were generated from basaltic precursors; and (iii) geomorphologic evidence that the Karoo province has experienced substantial permanent uplift associated with volcanism. Based on these observations, he estimates the presence of a ca. 5 km thick underplated gabbroic layer in southern Africa, but the few refraction seismic profiles and receiver functions from the region indicates that the Archaean crust is thin and that a high-velocity lower crust is lacking (Durrheim and Mooney, 1994; Youssof et al., this volume).

It is difficult to identify direct evidence for magmatic underplating below the cratons, as cratons often include high-velocity lower crust. These lower crustal layers may have been formed by underplating

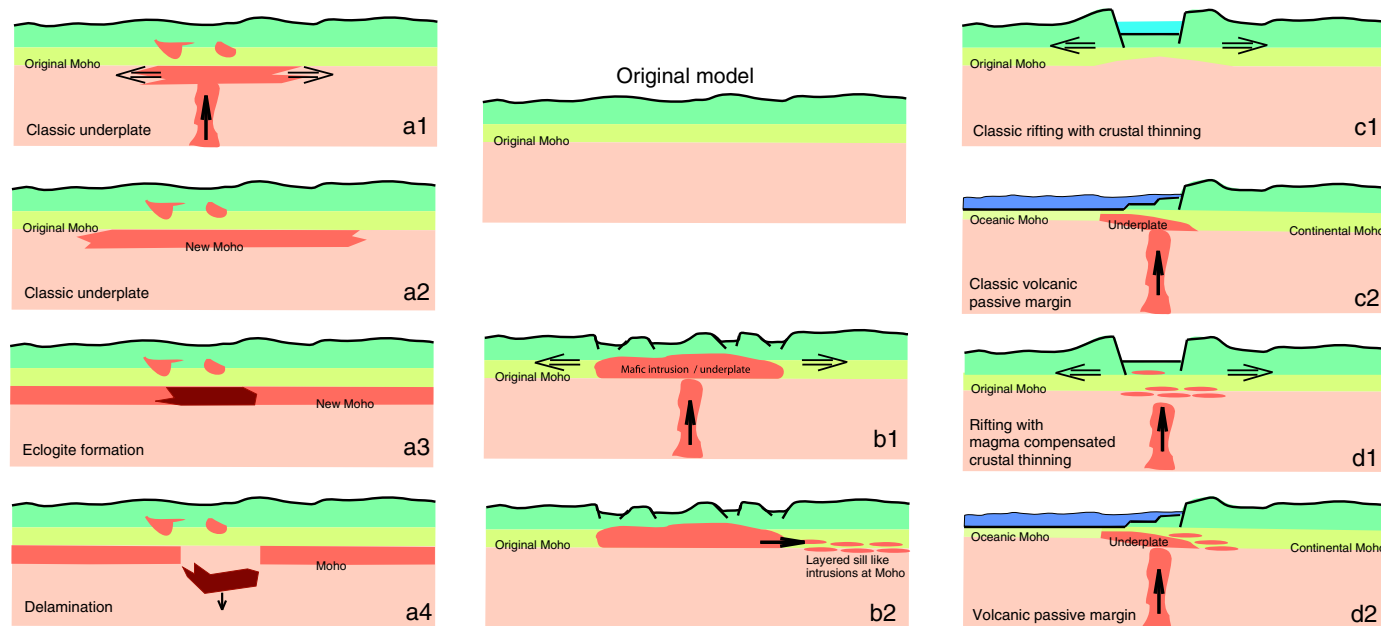


Fig. 13. Conceptual models of processes associated with underplating. Cases a–d are all based on the same original model of the crust and mantle. Case a: Classic underplating process (a1) leading to a large underplated layer below Moho (a2). Parts of such underplated layer may eventually be metamorphosed into eclogite facies (a3) and possibly be partially delaminated at a later stage (a4). Case b: Crustal stretching leading to massive mafic intrusion/underplating in the form of a crustal batholith (b1); and layered intrusion/underplate along the Moho, possibly at a late stage of development (b2). Case c: Classic passive rifting due to extension caused by far field stresses leading to uplifted Moho (c1); successful break-up may be accompanied by mafic intrusion/underplating at the continent–ocean boundary (c2). Case d: Magma compensated crustal thinning associated with rift development creating mafic intrusions/underplate in the lower crust and around Moho (d1); successful break-up may be accompanied by mafic intrusion/underplating at the continent–ocean boundary, with complicated resulting structure from the two phase development (d2).

processes but it is difficult to distinguish such formation process from other options. The Siberian craton shows increased lower crustal reflectivity in some places, but mainly around sedimentary grabens and basins which may indicate that they are caused by extension (Cherepanova et al., *this volume*).

Considering the presumed geodynamic importance of large igneous provinces and plumes, surprisingly few seismic profiles and datasets have been acquired in these environments. The Deccan Traps have some seismic coverage, contrary to most other LIP. Based on these images Reddy and Vijaya Rao (*in press*) interpret that an up-to 12 km thick layer was underplated during the formation of the Deccan Traps. It has been observed in five refraction seismic profiles, but no normal-incidence reflection sections exist. The underplated layer is highly seismically reflective and has high seismic velocity between 6.9 and 7.3 km/s.

The presumed plume at Yellowstone seems to have variable influence on the crustal velocities, such that some volcanoes are underlain by low-velocity bodies and other areas are underlain by high-velocity zones in the lower crust (Stachnik et al., 2008). The low velocities may be explained by high temperature below the presently active volcanoes, whereas the high velocities potentially could be caused by magmatic bodies of underplated type, but seismic observations indicate that underplating may not be a dominant process at Yellowstone.

6. Discussion

The presented examples of crustal underplating by mantle-derived basaltic material demonstrate that the processes related to magmatic intrusion at the base of the crust are highly variable, and that the resulting structure may take on many forms (Fig. 13). The early understanding of underplating processes inferred widespread ponding of magma just below the Moho as expected from petrologic considerations (e.g. Cox, 1980; Fyfe, 1978), but it is uncertain if structure related to such general underplating has ever been observed by geophysical imaging. Instead the available seismic models tend to image relatively localised structure of up-to 200 km in width, for example at magmatic island arcs and continental margins, where the underplated material is observed as a high-velocity lower crust. In both types of such locations, the observed high-velocity bodies are generally without internal seismic reflectivity. This may indicate a homogeneous composition and long duration of solidification after magma addition, during which time the magma fractionates into a light component that rises to the upper crust and the high-density residue above the Moho. Island arc processes may be a main source of new continental crust if the upper and lower magmatic bodies may efficiently be separated, and may lead to delamination of the underplated material. It may be speculated if the separation rather takes place by long-distance tectonic movement which, however, does not explain how the high-velocity lower crust may be returned to the mantle.

Underplated material has high density and therefore it is valid to assume that it also has high seismic velocity. Mafic rocks may be identified by a high Poisson's ratio (high V_p/V_s ratio) as compared to ultra-mafic rocks (Christensen, 1996). This provides possibilities for geophysical identification of underplated material, although there are relatively few S-wave studies available.

The underplated bodies usually do not show internal seismic structure but are clearly distinguished from the surrounding host rock by their high seismic velocity. However, their upper and lower boundaries usually constitute distinct seismic reflectors that are imaged as both normal-incidence and wide-angle reflections. Despite the relatively small velocity contrast at the Moho (from ca. 7.3 to 8.0 km/s) the Moho is usually also a distinct reflector, taken as the base of the underplated body. An exception has been observed in the central part of a magmatic intrusion in the Norwegian–Danish Basin (Fig. 5), where the latest cooling formed a rock with very high velocity and almost no contrast to the upper mantle velocity.

Seismically highly reflective underplated material is mainly observed below continental rift zones in up-to 15 km thick layers. The reflectivity can be explained by the presence of elongated thin mafic bodies (generally 300–500 m thick and km-scale long) which may be interpreted as sill-like intrusions in the pre-existing lower crust. These intrusions may fully or partially compensate the crustal thinning caused by lithosphere extension during rifting. In some cases (e.g. below the Oslo Graben) the underplated material appears non-reflective as it would be in the case of magma ponding at the elevating Moho. Some rift zones show no sign of magmatic addition to the lower crust/Moho depth region (e.g. the central Graben in the North Sea). Strong lateral variability of the Moho depth and the crustal reflectivity may exist along strike of rift zones (e.g. in the Kenya and Rhine Graben rift zones, Figs. 5 and 7).

New seismic profiling at the North Atlantic margin has demonstrated some seismic reflectivity in the high-velocity lower crustal zone which may be related to magmatic underplating at the continent to ocean transition. The reflectivity appears to be restricted to parts of the high-velocity body, and it may be rather related to the continental rifting that predated break-up, whereas massive intrusion of magma took place at the time of break-up.

Underplating was originally proposed in the 70's as a major crust forming process in cratonic settings around the large igneous provinces and other major magmatic provinces. Nevertheless, the existing geophysical data is very limited and, so far, inconclusive, although some new studies provide indication for the presence of major underplated structures in such settings. Surprisingly, underplating still has to be demonstrated as a feature of active orogens. It is possible that the high-velocity crustal roots observed in some cratons may originally have formed by underplating processes, but such generic relation yet has to be substantiated by data.

7. Conclusions

We have argued that the resulting structures of underplating processes may take many forms, and that the original concept of underplating, as widely distributed magmatic ponding below the Moho, never has been observed by geophysical data (Fig. 13). However, the idea that magma will pond where it achieves neutral buoyancy is a valuable concept and, as shown in this paper, geophysical data have provided evidence for structures that may represent underplated material in the sense that it has been formed from ponded magma.

Geophysical data and seismic images show that underplated material around the base of the crust is characterised by the following criteria (parameter ranges depend strongly on temperature):

1. High P-wave velocity in the range of 6.9–7.8 km/s. The highest values of 7.8 km/s usually mark a gradual boundary between the crust, as represented by underplated material, and the mantle.
2. High S-wave velocity, but the available data is sparse.
3. High V_p/V_s or Poisson's ratio as a discriminator for mafic material.
4. High density, which is characteristic for mafic material.
5. Seismic reflectivity which may be present or not, depending on the mode of intrusion of the underplated material. Strong seismic reflectivity is often observed where the intrusions have the form of sill like features, whereas reflection-free bodies may result from massive magma chamber cooling.

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